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Controls on the tropical response to abrupt climate changes.

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Abrupt climate change events during glacial times have distinct tropical imprints, despite their cause being at high latitudes. The mechanisms by which high latitudes affect low latitudes are currently unclear. We present climate model simulations of a set of different abrupt events and find changes in tropical rainfall can not be comprehensively explained by either changes in sea ice extent or ocean circulation. Changes in tropical meridional temperature gradients are the clearest way to explain tropical rainfall changes across all events. We find some tropical regions are unlikely to record Heinrich Events because the rainfall is moved far enough away from them during stadial periods that they are insensitive to further change. Greenland temperature responds linearly to Atlantic sea ice extent suggesting that the absence of change in Greenland temperature during Heinrich Events implies no sea ice expansion, despite major changes in the climate system elsewhere.

1. Introduction

There is a large body of evidence that the abrupt climate changes of the last glacial period so clearly seen in North Atlantic climate records can be seen much further afield [e.g. Deplazes et al., 2013]. Understanding these speaks to our understanding not only of the climate of the past, but also to the basic dynamics of the climate system.

There were two distinct abrupt climate changes in the North Atlantic during the last glacial period: Dansgaard/Oeschger (DO) Events and Heinrich Events. DO Events are most clearly seen in Greenland ice core records and are likely related to changes in North Atlantic sea ice [Gildor and Tziperman, 2003; Li, 2005; Dokken et al., 2013]; Heinrich Events occurred during the cold, stadial, periods of some DO Events and are not seen in Greenland ice core records. By definition Heinrich Events are related to collapses of Northern Hemisphere Ice Sheets and their attendant fluxes of water and debris into the North Atlantic [Heinrich, 1988; Hemming, 2004]. While it has been suggested that the ocean circulation is fundamental in explaining changes in North Atlantic climate [Broecker et al., 1985], its exact role is debated. During Heinrich Events there were undoubtedly fluxes of freshwater into the ocean which could have disrupted the circulation, but it is unclear whether any observed changes occurred after, before or even caused the ice sheet collapse [Alvarez-Solas and Ramstein, 2011; Marcott et al., 2011; Lynch-Stieglitz et al., 2014]. During DO Events there is evidence that the ocean circulation evolved from being relatively slow and shallow during the stadials to deeper and more vigourous during the interstadials [e.g. Burckel et al., 2015; Henry et al., 2016], however, this evidence is not conclusive [e.g Them et al., 2015; Thornalley et al., 2013]. Furthermore, it is not obvious whether the ocean circulation is a unique driver of change or just part of an internal oscillation of the coupled ocean-atmosphere-sea ice system [Drijfhout et al., 2013; Martin et al., 2015; Sidorenko et al., 2014; Kleppin et al., 2015; Vettoretti and Peltier, 2018; Brown and Galbraith, 2016].

Changes in the location of Tropical rainfall during Heinrich and DO Events have been linked to changes in the high latitudes through a number of mechanisms. Chiang and Bitz [2005] showed that cooling at high northern latitudes can propagate south, leading to colder sea surface temperature (SST) north of the Equator. These act to move the Inter Tropical Convergence Zone (ITCZ) southward [the importance of local tropical SST changes in this movement was further investigated by Cvijanovic and Chiang, 2012]. Marshall et al. [2013] and Schneider et al. [2014] argued that because the ITCZ and the location of ascent in the Hadley Circulation are linked, and because the Hadley Circulation is responsible for much of the heat transported between hemispheres by the atmosphere, an ITCZ shift can be linked to changes in atmospheric heat transport. Through this relationship it has been argued that the ITCZ adopts its position to maintain the global energy balance. Indeed Marshall et al. [2013] argue that the mean position of the ITCZ north of the equator, and implied southward energy transport, is required to balance the ocean's northward heat transport. Anything that acts to perturb the global energy balance, or heat transported by the ocean, could, therefore, directly lead to a movement in the ITCZ. Since northward heat transport in the ocean is strongly related to the Atlantic Meridional Overturning Circulation (AMOC), it has been proposed that AMOC changes can be related to movements of the ITCZ [Mulitza et al., 2017]: AMOC weakening will tend to cool the NH, with an associated southward movement of the ITCZ. More recent work has suggested that wind-driven ocean dynamics can weaken the sensitivity of the ITCZ to extratopical heating, however [Schneider, 2017; Green and Marshall, 2017; Kang et al., 2018].

These two mechanisms offer a way to understand why the ITCZ moves during abrupt climate changes whose root is in the high latitudes. One view indicates a need to understand the processes by which the high latitudes can alter the surface climate in the tropics; the other indicates that we need to understand how energy transports evolve. The second, energy based view, potentially links ocean circulation changes, which will change how the ocean transports heat, to changes in the location of the ITCZ. In particular, it suggests that we ought to be able to predict how much the ITCZ shifts given how much the ocean circulation changes [Mulitza et al., 2017]. We shall address this question here: is there a general mechanism that can explain why tropical rainfall responds to abrupt events in high latitudes?

A general description of the impact of DO and Heinrich Events on the tropics must answer two further questions. First, since Greenland temperatures were no different during Heinrich Events to the preceding DO stadial, it must explain why the ITCZ moves south despite the North Atlantic apparently being no cooler during Heinrich Events. Second, while many locations in the Tropics indicate progressive drying from DO stadials into Heinrich Events, others do not. The Cariaco Basin reflectance record [Peterson, 2000] is an example of this. Why, therefore, do some tropical records appear insensitive to Heinrich Events?

In this study we shall compare the tropical precipitation response of the climate to DO and Heinrich Events by imposing two different types of forcing to a coupled atmosphere/ocean climate model. To simulate DO stadials we impose a sea ice concentration in the North Atlantic and Nordic seas; to simulate Heinrich stadials we impose a hosing to the North Atlantic. In Nature no two DO or Heinrich Events are the same. To understand how different events might affect the climate we also vary size of the forcing. Understanding the trends in the climate response allows us to understand the differences between events and, more importantly, draw out any general relationships about ITCZ movements.

2. Methods

In order to simulate the effects of abrupt changes there has been some success using "hosing" simulations: model simulations in which an artificial freshwater flux is introduced into the ocean in order to disrupt the AMOC [Manabe and Stouffer, 1995]. While this may be a suitable model set up for simulating Heinrich Events, in which there is a flux of freshwater, it is not clear that this is representative of DO Events. For, while hosing simulations increase the sea ice concentration in the Nordic Seas and North Atlantic in agreement with proxy data, they also disrupt the ocean circulation in a very particular way [Chang et al., 2008]. Since it is not clear that there was a significant source of freshwater during DO Events, it is not clear that the distinct ocean circulation changes simulated by hosing simulations will match those seen during DO Events. Some models do spontaneously produce DO like Events [Drijfhout et al., 2013; Sidorenko et al., 2014; Kleppin et al., 2015; Vettoretti and Peltier, 2018; Brown and Galbraith, 2016], due to feedbacks between sea ice cover and ocean stratification, but it is not clear how similar the ocean responses are in these simulations to those with hosing. It is important to understand these differences, because outside the North Atlantic the climate will respond more strongly to ocean circulation changes than to increases in sea ice.

Here, we use the fully coupled climate model HadCM3 [Gordon et al., 2000; Pope et al., 2000; Valdes et al., 2017]. Although this model does a good job at simulating tropical rainfall it is known to have too little cloud cover that is too optically bright [e.g. Massey et al., 2015]. This is a common deficiency in GCMs, including nearly all of those in CMIP5 [Nam et al., 2012]. Low clouds are known to be important in determining tropical rainfall [Hwang and Frierson, 2013].

Our hosing simulations impose a freshwater flux uniformly across the North Atlantic between 50-70°N. Very similar results can be obtained using a freshwater distribution that better simulates freshwater released from a collapse of the Laurentide Ice Sheet [see supplement Roberts et al., 2014]. We choose to use the uniform forcing, however, for better comparability with other studies. We vary the forcing between 0.04–1.0 Sv($10^6 m^3 s^{-1}$). To impose an annual mean sea ice cover to the north Atlantic we modify the sea-ice module in the model, artificially growing sea ice to ensure that sea-ice persists at a specified location and to a specified depth and concentration. We vary the location of the southern edge of this annual mean sea ice to be between 70–40°N. The sea ice depth is 4m and has a concentration of up to 95%. Note that because sea ice can be advected in the model there are changes to the sea ice distribution outside the area in which sea ice is imposed. We also undertake a set of simulations in which we prescribe a sea ice concentration and also impose a hosing. With these simulations we can better deconvolve how the climate responds to both sea ice and hosing. They are also the best analog for Heinrich Events as they account for both extended sea ice typical of stadial periods and freshwater fluxes from the Laurentide Ice Sheet.

Neither hosing simulations nor the sea ice specified simulations are strictly conservative: in both types of simulation energy is not conserved (in hosing simulations mass is not conserved; in sea ice simulations the energy budget is slightly perturbed). The theory relating the ITCZ and energy transports requires a closed energy budget. However, the relationship is still present, and indeed robust, in models in which the energy budget is far from closed [Donohoe et al., 2013; Roberts et al., 2017].

DO and Heinrich Events did not occur throughout the last glacial period which suggests the occurrence of these events may depend upon the background climate state. To investigate this possible state dependence we undertake two sets of simulations: one perturbing about an LGM climate and one about a pre-industrial.

Because we use a coupled climate model, both the oceanic and atmospheric circulations can respond to forcing. We find that under both forcing regimes there is a weakening of the AMOC and a shift in the location of the ITCZ. It is to these changes that we now turn our attention.

3. The relationship between the ITCZ and the ocean or sea ice.

We examine ITCZ location using the centroid of global precipitation [P_{cent} Donohoe et al., 2013]. This is a zonal mean metric which emphasises the relationship between energy transports and the ITCZ. We caution, however, that local rainfall changes can be quite unrelated to the zonal mean [Roberts et al., 2017; Singarayer et al., 2017]. Other ITCZ location metrics exist, however P_{cent} best emphasises the relationship between atmospheric heat transport and ITCZ location [Roberts et al., 2017].

Weakening of the AMOC can reduce the northward ocean heat transport. This could alter the ITCZ location through some energetic constraint [Marshall et al., 2013] or because, as the North Atlantic cools, the ITCZ moves [Chiang and Bitz, 2005]. We first examine the relationship between ITCZ location and AMOC strength. Our experiments show a general trend for a southward ITCZ shift with weaker AMOC (Figure 1(a)). This is most apparent in the set of hosing simulations (blue squares). However, there is a large spread in the response – a 0.3° southward ITCZ shift is related to a change in AMOC strength of anything between 0 and -12 Sv. If AMOC changes were a dominant control on ITCZ location we would not expect to see such a spread.

While AMOC strength is doubtless related to northward ocean heat transport, changes in the temperature structure of the ocean or the wind-driven surface circulation could also change ocean transport [Schneider, 2017; Green and Marshall, 2017; Kang et al., 2018]. To more directly measure the link between the ITCZ and the ocean we directly calculate the ocean heat transport. Although this may be calculated within a model, this is unknown in the palaeo-ocean. Our simulations show a general southward ITCZ movement as ocean heat transport decreases: this relationship is not robust, however (Fig. 1(b)). Hosing simulations using a pre-industrial climate show the strongest relationship (blue squares), with other experiments showing lower correlation. It is notable that experiments in which we prescribe sea ice (circles) show quite large changes in ITCZ location for no changes in ocean heat transport. From this we conclude that it is not possible to make a general relationship between ITCZ position and ocean transports. The specific case of hosing in a pre-industrial climate (blue squares) demonstrates a strong linear relationship between both the strength of the AMOC, northward ocean heat transport and the position of the ITCZ. This relationship, however, is unique to this one experiment.

It has been suggested that the ITCZ will move southward as sea ice extent in the North Atlantic increases [Chiang and Bitz, 2005; Rhodes et al., 2015]. We find that as sea ice area increases the ITCZ does move southwards; however this is not a strong relationship (Fig. 1(c)). Amongst hosing simulations (blue squares) there are significant movements in the ITCZ when the amount of hosing increases despite no increase in the sea ice extent. Furthermore, comparing the experiments in which we impose sea ice and include a hosing (coloured markers surrounded by square or circle), we find changes in ITCZ location despite the same sea ice distribution. We thus conclude that the ITCZ location does not respond directly to sea ice area in the North Atlantic. We should note that in the suite of experiments in which we prescribe the amount of sea ice (black filled circles), there is a strong linear relationship between sea ice are and ITCZ location. However, this relationship is unique to this experimental set up.

We have demonstrated that it is not possible to generalise the relationship between the ITCZ location and either the state of the ocean circulation, or the amount of sea ice in the Northern Hemisphere. This begs the question: is there some mechanism that can explain the shift in the ITCZ regardless of how the model is forced? In common with other studies [Donohoe et al., 2013; Roberts et al., 2017] we do find a strong relationship between the atmospheric heat transport across the equator and ITCZ location (Fig. 1(f)). However, this is not a practical way to explain tropical rainfall changes because we have no way of predicting how the equatorial heat transport will change in response to the extratropical forcings that are imposed.

Experiments in which we impose both hosing and a sea ice distribution offer a way to assess a general relationship. By prescribing sea ice area in the North Atlantic, the high latitude temperature change remains the same, eliminating this as a cause; we have just shown that ocean circulation changes are not a direct cause either.

4. How the ITCZ is related to abrupt changes.

We find in all simulations where hosing is imposed in addition to a specified sea ice extent, there is a southward ITCZ movement (see the markers in Fig. 1 outlined by the open square or circle). Furthermore, the ITCZ is located farther south when the amount of freshwater introduced is larger. For one particular case [sea ice imposed to 50°N in the Atlantic and 0.25Sv freshwater forcing] we see a large change in the two dimensional rainfall distribution when the hosing is introduced in addition to the sea ice; this accounts for the southward shift of the zonal mean ITCZ (Fig. 2(a)). As expected there is negligible change in the sea ice distribution (Supplement). Maps of surface temperature change show that there is no temperature change at high latitudes in the North Atlantic (Fig 2(b)). Looking near the equator, we see that there is a distinct cooling extending along the eastern and southern edge of the Atlantic subtropical gyre, giving a broad cooling at 15°N. South of the equator we see a more muted, but equally widespread, warming. This pattern is very reminiscent of the meridional mode that was originally proposed as the cause of the ITCZ shifts [Chiang and Bitz, 2005].

To generalise the response we compute an EOF of tropical precipitation anomalies across all of the simulations. The pattern that explains most of the variance in precipitation (75%) is a movement to the south (Fig. 2(c)). It is notable that the largest changes in the rainfall are in the Atlantic extending into the far East Pacific. Here the pattern is similar to a double ITCZ bias. This pattern is the same if we use only the sea ice forced or the hosing simulations (Supplement). The surface temperature field that is associated with this rainfall pattern shows a distinct cooling around 15°N in the Atlantic (Fig. 2(d)). This is very similar to the pattern shown in the one single pair of experiments imposing sea ice with and without hosing (Fig.2(b)).

In order to generalise the zonal mean ITCZ response we show in Fig. 1(d) P_{cent} plotted against the temperature difference across the equator (the difference between 10°S-10°N and 10°N-30°N). This is a similar metric to that used in Donohoe et al. [2013] but more closely reflects the region shown in Fig. 2(d) to be associated with the changing rainfall patterns. We find a very close relationship. Unlike all of the other metrics examined, this equatorial temperature difference can explain movements in the ITCZ for all experiments, furthermore it holds for both a pre-industrial and LGM background state. We contrast this with plotting P_{cent} against the temperature difference between the high northern latitudes and the equator (Fig. 1(e)), in which case we find little relationship.

All of our analysis has focused on the zonal mean response of the ITCZ. However Figure 2 shows that both the rainfall and surface temperature changes are not zonally homogeneous. The largest responses are centred around the Atlantic. To understand how much of the zonal mean response arises from the Atlantic we recalculate P_{cent} over the Atlantic and Eastern Pacific (240°E – 360°E) basin and the remainder of the globe (0°E – 240°E). We find that P_{cent} in the Atlantic and the zonal mean are highly correlated, although the Atlantic P_{cent} changes by in excess of 7° in comparison to zonal mean changes of only 3.5° (Fig. 3(a)). By contrast, there is little correlation between the zonal mean ITCZ and the ITCZ outside of the Atlantic (Fig. 3(b)). Comparing the meridional temperature gradient and P_{cent} in the two regions we find the same result: there is a strong relationship in the Atlantic but no relationship outside this region (Fig. 3 (c) and (d)). We thus conclude that the relationships we have shown in the zonal mean quantities are driven by changes in the Atlantic. This is understandable because the forcing that we impose is in the Atlantic.

5. Implications.

We have shown that it is not possible to explain changes in tropical rainfall in terms of only changes in the ocean circulation or the area of sea ice in the North Atlantic. Rather, we must understand the changes in the temperature gradient within the tropics themselves. This can explain how we might expect to see impacts from DO and Heinrich Events in the tropics. In the Introduction we described two features of Heinrich and DO Events that must also be explained.

First Heinrich Events do not appear distinct in the Greenland temperature record yet there are changes in the tropical rainfall. Our model simulations show that the temperature in central Greenland is strongly correlated with the area of sea ice in the adjacent north Atlantic (Fig. 4(a)) in agreement with previous studies [Li, 2005]. This linear relationship continues even with the huge increase in sea ice area that occurs when the sea ice edge is imposed to be at 40°N. We can infer from this that because there is no evidence of temperature changes during Heinrich Events in Greenland, there was no increase in sea ice extent in the North Atlantic during these events. We find no relationship encapsulating all of the experiments between the strength of the AMOC and Greenland temperature (Fig. 4(b)) which indicates that there is no direct influence of changes in the ocean circulation on Greenland temperature.

In the tropics we do see evidence of both Heinrich and DO Events. We have argued that the tropical response to abrupt climate changes is caused by a change in the meridional gradient of temperature near the equator. Our simulations show this can be caused by both sea ice expansion and hosing. Therefore we conclude that during DO stadials there is extended sea ice which cools both Greenland and the northern tropical Atlantic with the latter cooling acting to move the ITCZ south. During Heinrich Events, which all occur during DO stadials, the freshwater flux does not expand sea ice any further, thus there is no change in the temperature of Greenland, but it does change the temperature in the northern tropical Atlantic, moving the ITCZ even further southward. This can be seen in our simulations with both sea ice and hosing imposed (Fig. 1)). Brown and Galbraith [2016] showed similar results.

The second feature is the puzzling absence of Heinrich Events from some tropical records. The highly time resolved Cariaco Basin record is an example of this [Peterson, 2000]. Examining rainfall in the catchment of the Cariaco Basin in our simulations we find that as the ITCZ moves southward, the rainfall amount falls almost linearly until the ITCZ reaches around 2.5°S (Fig. 4(c)). If the ITCZ moves farther south than this, the rainfall amount does not change because it is no longer directly caused by rainfall in the ITCZ. This is most clearly seen if we only consider the ITCZ in the Atlantic Basin, in which case when the ITCZ has reached 6°S the rainfall amount does not change (Fig. 4(d)). Thus we do not see Heinrich Events over Cariaco Basin because the ITCZ is already far enough south during the DO stadial that it is insensitive to any further southward movements forced by a Heinrich Event. We can contrast this with rainfall over the Arabian Sea. In this region rainfall continues to decrease as the ITCZ moves south during DO stadials into Heinrich Events (Fig. 4(e)).

6. Concluding remarks

We have shown that it is not possible to directly link changes in the ITCZ location to either changes in the ocean circulation or the area of sea ice. Instead, we must understand local temperature changes within the Tropics. Both ocean circulation changes and expansion of sea ice can cause changes in the high latitude energy budget, but we do not find that these changes are simply balanced by a cross equatorial heat flux. Although there is a strong link between the location of the ITCZ and atmospheric heat transport, changing the heat transport is but one way that the climate can balance a change in the energy budget. Many studies have shown that local "compensation" is an effective way to balance energy budget changes: our results echo this [Cvijanovic and Chiang, 2012; Kang et al., 2009]. This compensation can also reside in the ocean [Schneider, 2017; Green and Marshall, 2017; Kang et al., 2018]. This offers an explanation for why we see no clear relationship between the imposed changes in ocean circulation and the climate response: part of the climate's response involves further changes in the ocean. This further reinforces our result that there is no *simple* relationship that can be applied to data sparse past climates.

Although we have shown that the surface temperature gradient is the clearest way to understand the changes in the ITCZ location, we are not able to explain what causes the changes to the surface temperature. Chiang and Bitz [2005] suggest that tropical temperatures are altered through a series of feedbacks between the ocean and atmosphere: Schneider et al. [2014] and Bischoff and Schneider [2014] suggest they are linked to changes in the energy budget. In the latter case our analysis suggests that these changes are not simple and likely highly non-linear.

Our hosing simulations could be interpreted as demonstrating a link between the AMOC and the location of the ITCZ. However by using an alternative experimental set up we demonstrate that this link is merely because hosing simulations cause both a change in the AMOC and a change in the equatorial surface temperature gradient, with the ITCZ location responding to the latter change.

We thus end with the caution that although hosing simulations are a way to perturb the climate, they do alter it in a very particular way. Therefore, it is of utmost importance that when using the results of a hosing simulation to interpret an abrupt change in a climate proxy record, the root cause of the abrupt change is actually caused by something resembling a hosing simulation. Until such time as unequivocal ice-raft debris records are found to suggest that DO events arose from large injections of freshwater to the ocean, this precludes hosing simulations as being a useful way to interpret DO events.

7. Acknowledgments

Model output used in this manuscript may be accessed from https://www.paleo.bristol.ac.uk. POH is supported by a University of Birmingham fellowship.

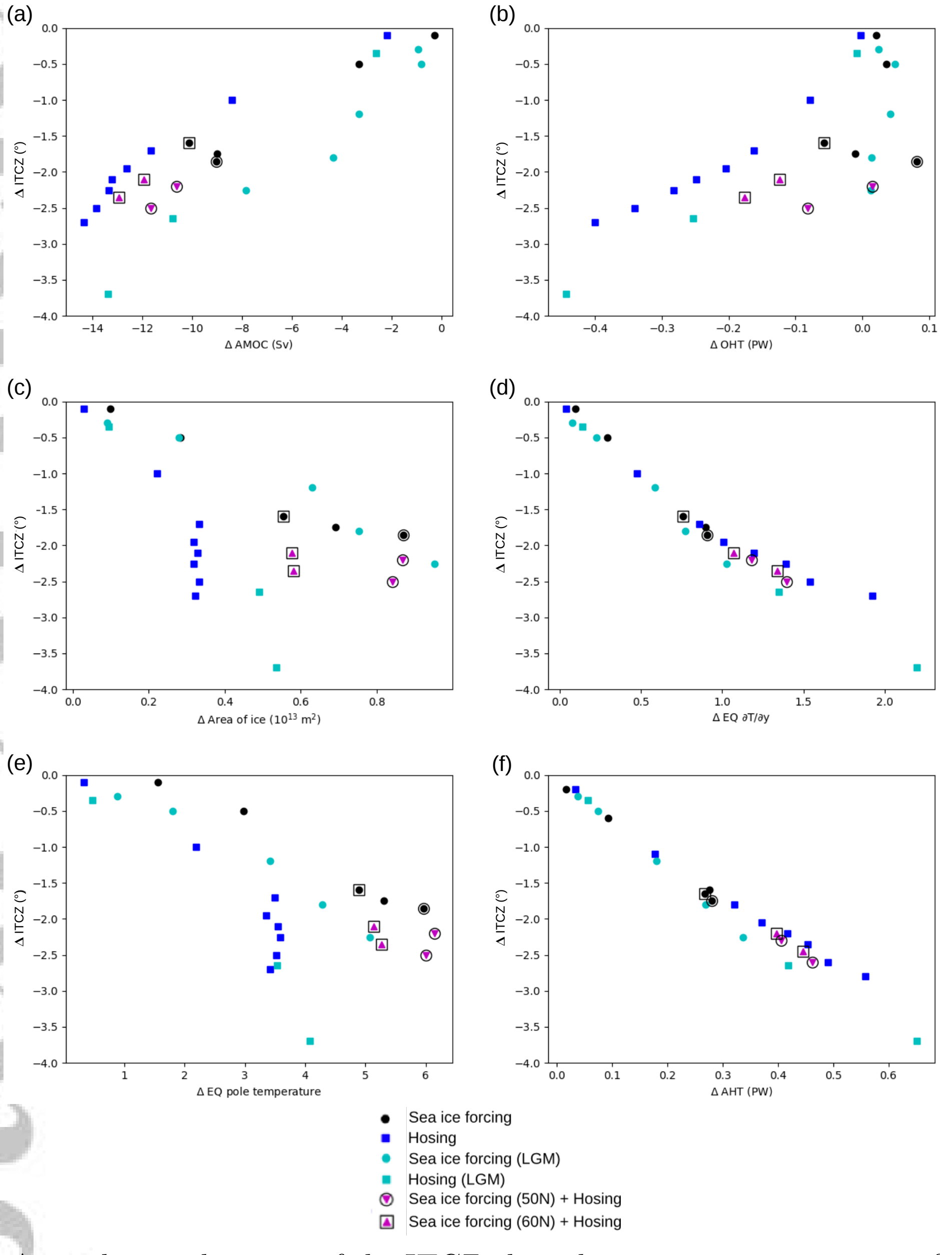


Figure 1. Annual mean location of the ITCZ plotted against various metrics of the climate for all simulations. All are plotted as anomalies relative to either a pre-industrial or LGM control. (a) ITCZ v. AMOC (defined as the maximum value of the overturning streamfunction in the Atlantic ocean north of 30°N). (b) ITCZ v. ocean heat transport across the equator computed explicitly. (c) ITCZ v. sea ice area in the North Atlantic. (d) ITCZ v. meridional surface temperature gradient in the northern tropics (average temperature difference between (10°S-10°N and 10°N-30°N). (e) ITCZ v equator to pole temperature difference in the NH (average temperature difference between 10°S-10°N and 60°N-80°N). (f) ITCZ v atmospheric energy transport across the equator.

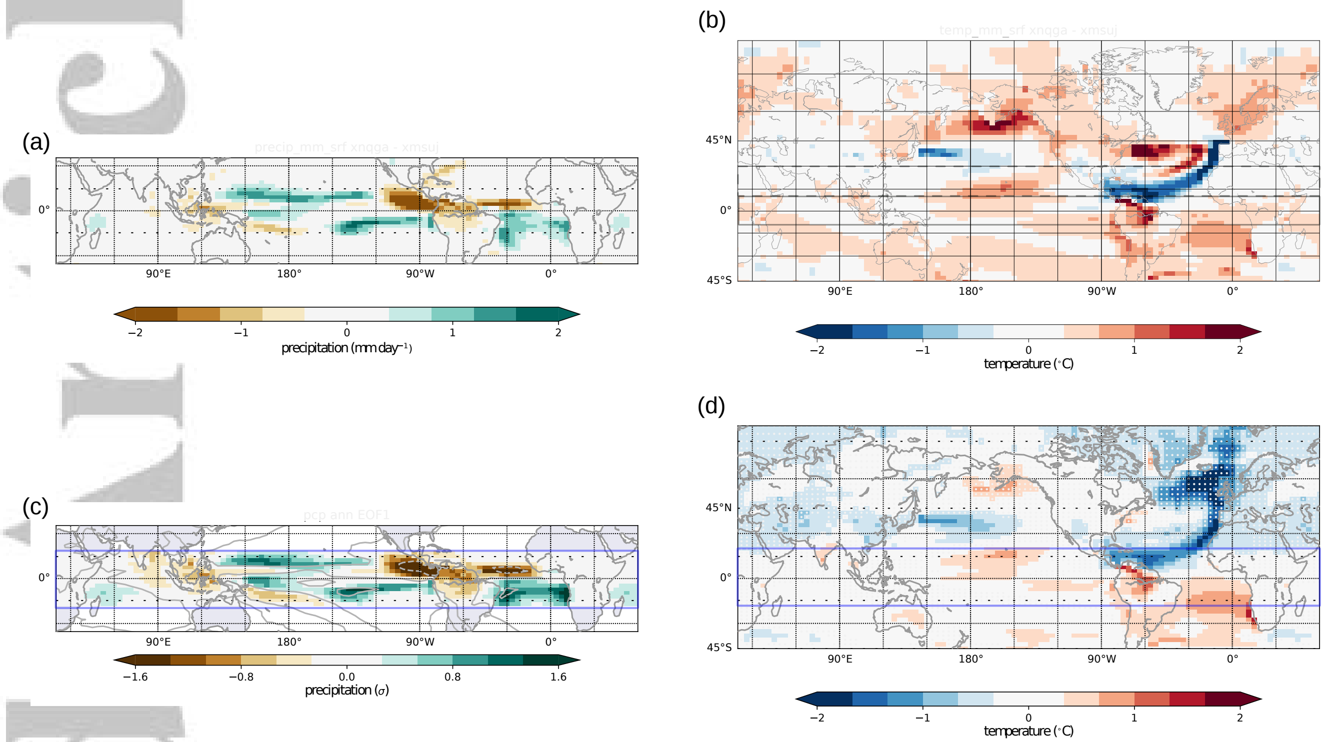


Figure 2. Maps showing climate response to hosing and imposed sea ice. (a) The difference in precipitation between a sea ice imposed simulation and a sea ice and hosing imposed simulation. (b) The difference in surface temperature between a sea ice imposed simulation and a sea ice and hosing imposed simulation. (c) First EOF of change in tropical (20 °N – 20°S) precipitation across all simulations. (d) Surface temperature and wind pattern associated with the precipitation pattern shown in (c). The blue outline box in (c) and (d) indicates the area over which the EOF of precipitation is calculated. The grey boxes in (b) indicate (10°S-10°N and 10°N-30°N) the area over which we calculate the tropical/subtropical temperature in Fig.1 (d)

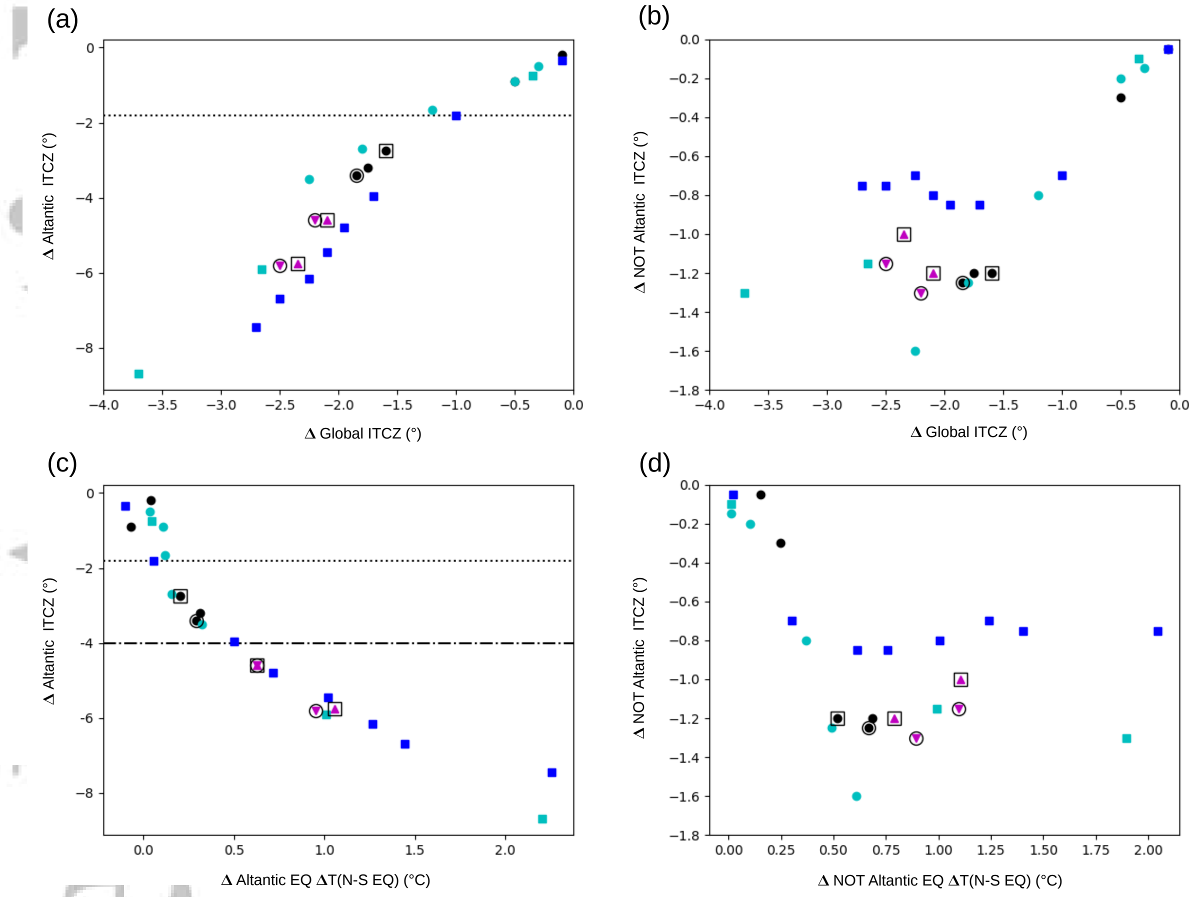


Figure 3. Relationship between Global ITCZ and the Atlantic. (a) Change in the location of the Atlantic (240°E-0°E) ITCZ plotted against the global ITCZ. (b) Change in the location of the ITCZ everywhere but the Atlantic (0°E–240°E) plotted against the global ITCZ. (c) Change in the location of the Atlantic ITCZ plotted against Atlantic meridional surface temperature gradient in the Northern Tropics. (d) Change in the location of the ITCZ everywhere but the Atlantic plotted against meridional surface temperature gradient in the Northern Tropics everywhere but the Atlantic.

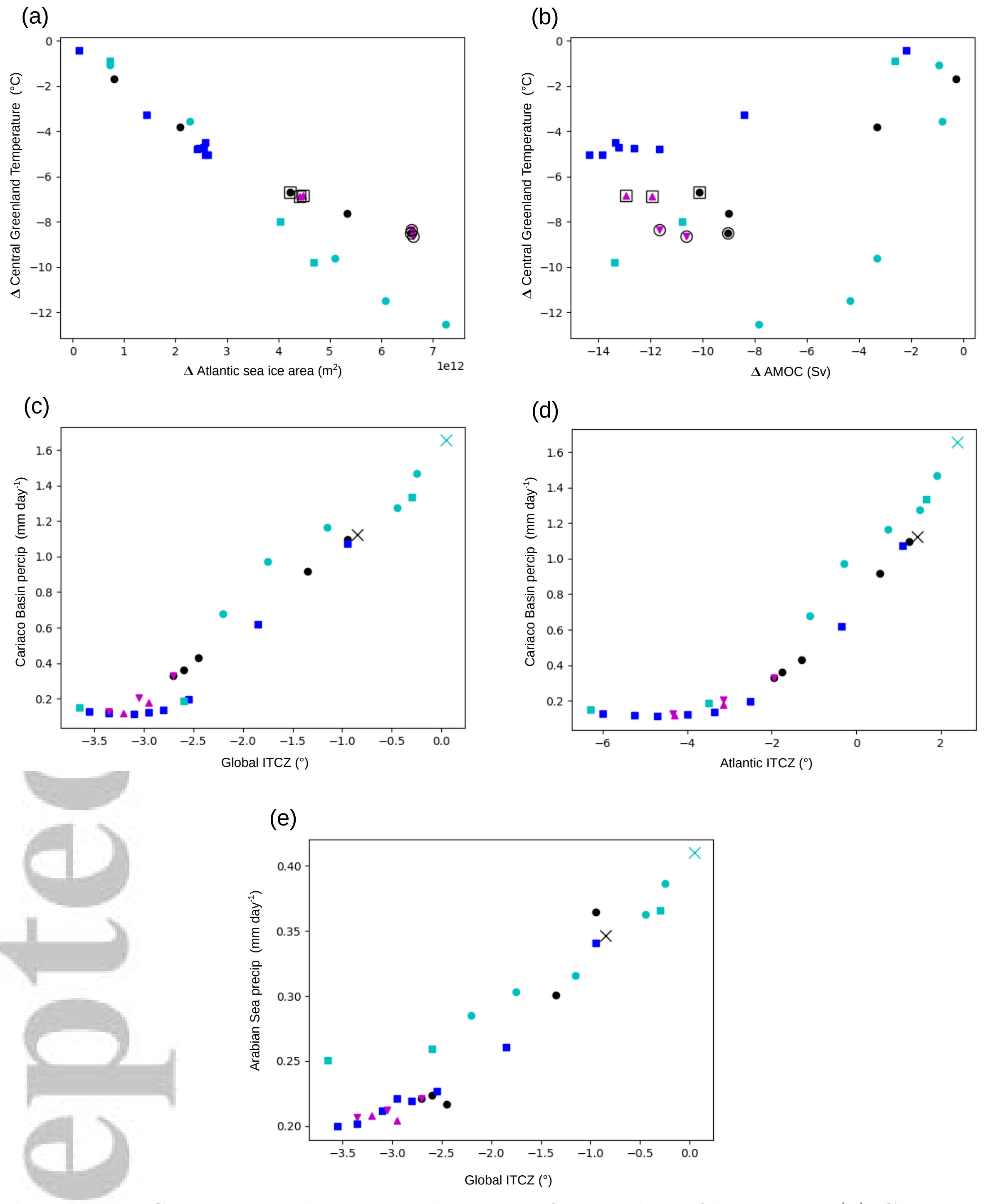


Figure 4. Climate variables at the location of a number of locations. (a) Change in central Greenland temperature plotted against change in Atlantic sea ice area. (b) Change in central Greenland temperature plotted against change in AMOC strength. (c) Total rainfall over Cariaco Basin catchment plotted against global ITCZ location. (d) Total rainfall over Cariaco Basin catchment plotted against Atlantic ITCZ location (P_{cent} calculated over 240°E to 0°E.) (e) Total rainfall over the Arabian sea plotted against global ITCZ.

References

Alvarez-Solas, J., Ramstein, G., 2011. On the Triggering Mechanism of Heinrich Events. Proceedings of the National Academy of Sciences 108, E1359–E1360. doi:10.1073/pnas.1116575108.

Bischoff, T., Schneider, T., 2014. Energetic Constraints on the Position of the Intertropical Convergence Zone. Journal of Climate 27, 4937–4951. doi:10.1175/JCLI-D-13-00650.1.

Broecker, W.S., Peteet, D.M., Rind, D., 1985. Does the ocean–atmosphere system have more than one stable mode of operation? Nature 315, 21–26. doi:10.1038/315021a0.

Brown, N., Galbraith, E.D., 2016. Hosed vs. unhosed: interruptions of the Atlantic Meridional Overturning Circulation in a global coupled model, with and without freshwater forcing. Climate of the Past 12, 1663–1679. doi:10.5194/cp-12-1663-2016.

Burckel, P., Waelbroeck, C., Gherardi, J.M., Pichat, S., Arz, H., Lippold, J., Dokken, T., Thil, F., 2015. Atlantic Ocean circulation changes preceded millennial tropical South America rainfall events during the last glacial. Geophysical Research Letters 42, 411–418. doi:10.1002/2014GL062512.

Chang, P., Zhang, R., Hazeleger, W., Wen, C., Wan, X., Ji, L., Haarsma, R.J., Breugem, W.P., Seidel, H., 2008. Oceanic link between abrupt changes in the north Atlantic ocean and the African monsoon. Nature Geoscience 1, 444–448. doi:10.1038/ngeo218.

Chiang, J.C.H., Bitz, C.M., 2005. Influence of high latitude ice cover on the marine Intertropical Convergence Zone. Climate Dynamics 25, 477–496. doi:10.1007/s00382-005-0040-5.

Cvijanovic, I., Chiang, J.C.H., 2012. Global energy budget changes to high latitude North Atlantic cooling and the tropical ITCZ response. Climate Dynamics 40, 1435–1452. doi:10.1007/s00382-012-1482-1.

Deplazes, G., Lückge, A., Peterson, L.C., Timmermann, A., Hamann, Y., Hughen, K.a., Röhl, U., Laj, C., Cane, M.a., Sigman, D.M., Haug, G.H., 2013. Links between tropical rainfall and North Atlantic climate during the last glacial period. Nature Geoscience 6, 1–5. doi:10.1038/ngeo1712.

Dokken, T.M., Nisancioglu, K.H., Li, C., Battisti, D.S., Kissel, C., 2013. Dansgaard-Oeschger cycles: Interactions between ocean and sea ice intrinsic to the Nordic seas. Paleoceanography 28, 491–502. doi:10.1002/palo.20042.

Donohoe, A., Marshall, J., Ferreira, D., Mcgee, D., 2013. The Relationship between ITCZ Location and Cross-Equatorial Atmospheric Heat Transport: From the Seasonal Cycle to the Last Glacial Maximum. Journal of Climate 26, 3597–3618. doi:10.1175/JCLI-D-12-00467.1.

Drijfhout, S., Gleeson, E., Dijkstra, H.A., Livina, V., 2013. Spontaneous abrupt climate change due to an atmospheric blocking-sea-ice-ocean feedback in an unforced climate model simulation. Proceedings of the National Academy of Sciences 110, 19713–19718. doi:10.1007/BF00391760.

Gildor, H., Tziperman, E., 2003. Sea-ice switches and abrupt climate change. Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences 361, 1935–1944. doi:10.1098/rsta.2003.1244.

Gordon, C., Cooper, C., Senior, C.A., Banks, H., Gregory, J.M., Johns, T.C., Mitchell, J.F.B., Wood, R.A., 2000. The simulation of SST, sea ice extents and ocean heat transports in a version of the Hadley Centre coupled model without flux adjustments. Climate Dynamics 16, 147–168. doi:10.1007/s003820050010.

Green, B., Marshall, J., 2017. Coupling of trade winds with ocean circulation damps itcz shifts. Journal of Climate 30, 4395–4411. doi:10.1175/JCLI-D-16-0818.1.

Heinrich, H., 1988. Origin and consequences of cyclic ice rafting in the northeast atlantic-ocean during the past 130,000 years. Quaternary Research 29, 142–152.

Hemming, S.R., 2004. Heinrich events: Massive late Pleistocene detritus layers of the North Atlantic and their global climate imprint. Reviews of Geophysics 42, RG1005. doi:10.1029/2003RG000128.

Henry, L.G., McManus, J.F., Curry, W.B., Roberts, N.L., Piotrowski, A.M., Keigwin, L.D., 2016. North Atlantic ocean circulation and abrupt climate change during the last glaciation. Science 353, 470–474. doi:10.1126/science.aaf5529.

Hwang, Y.T., Frierson, D.M.W., 2013. Link between the double-intertropical convergence zone problem and cloud biases over the southern ocean. Proceedings of the National Academy of Sciences 110, 4935–4940. doi:10.1073/pnas.1213302110.

Kang, S.M., Frierson, D.M.W., Held, I.M., 2009. The Tropical Response to Extratropical Thermal Forcing in an Idealized GCM: The Importance of Radiative Feedbacks and Convective Parameterization. Journal of the Atmospheric Sciences 66, 2812–2827. doi:10.1175/2009JAS2924.1.

Kang, S.M., Shin, Y., Xie, S.P., 2018. Extratropical forcing and tropical rainfall distribution: energetics framework and ocean Ekman advection. npj Climate and Atmospheric Science 1, 1–10. doi:10.1038/s41612-017-0004-6.

Kleppin, H., Jochum, M., Otto-Bliesner, B., Shields, C.A., Yeager, S., 2015. Stochastic atmospheric forcing as a cause of Greenland climate transitions. Journal of Climate 28, 7741–7763. doi:10.1175/JCLI-D-14-00728.1.

Li, C., 2005. Abrupt climate shifts in Greenland due to displacements of the sea ice edge. Geophysical Research Letters 32, L19702. doi:10.1029/2005GL023492.

Lynch-Stieglitz, J., Schmidt, M.W., Gene Henry, L., Curry, W.B., Skinner, L.C., Mulitza, S., Zhang, R., Chang, P., 2014. Muted change in Atlantic overturning circulation over some glacial-aged Heinrich events. Nature Geoscience 7, 144–150. doi:10.1038/ngeo2045.

Manabe, S., Stouffer, R.J., 1995. Simulation of abrupt climate change induced by freshwater input to the north atlantic ocean. Nature 378, 165–167. doi:10.1038/378165a0.

Marcott, S.a., Clark, P.U., Padman, L., Klinkhammer, G.P., Springer, S.R., Liu, Z., Otto-Bliesner, B.L., Carlson, A.E., Ungerer, A., Padman, J., He, F., Cheng, J., Schmittner, A., 2011. Ice-shelf collapse from subsurface warming as a trigger for Heinrich events. Proceedings of the National Academy of Sciences of the United States of America 108, 13415–9. doi:10.1073/pnas.1104772108.

Marshall, J., Donohoe, a., Ferreira, D., McGee, D., 2013. The ocean's role in setting the mean position of the Inter-Tropical Convergence Zone. Climate Dynamics 42, 1967–1979. doi:10.1007/s00382-013-1767-z.

Martin, T., Park, W., Latif, M., 2015. Southern Ocean forcing of the North Atlantic at multi-centennial time scales in the Kiel Climate Model. Deep-Sea Research Part II: Topical Studies in Oceanography 114, 39–48. doi:10.1016/j.dsr2.2014.01.018.

Massey, N., Jones, R., Otto, F., et al., 2015. weather@home – development and validation of a very large ensemble modelling system for probabilistic event attribution . Q J R Meteorol Soc 14, 1528–1545. doi:10.1002/qj.2455.

Mulitza, S., Chiessi, C.M., Schefuß, E., Lippold, J., Wichmann, D., Antz, B., Mackensen, A., Paul, A., Prange, M., Rehfeld, K., Werner, M., Bickert, T., Frank, N., Kuhnert, H., Lynch-Stieglitz, J., Portilho-Ramos, R.C., Sawakuchi, A.O., Schulz, M., Schwenk, T., Tiedemann, R., Vahlenkamp, M., Zhang, Y., 2017. Synchronous and proportional deglacial changes in Atlantic meridional overturning and northeast Brazilian precipitation. Paleoceanography 32, 622–633. doi:10.1002/2017PA003084.

Nam, C., Bony, S., Dufresne, J.L., Chepfer, H., 2012. The 'too few, too bright' tropical low-cloud problem in cmip5 models. Geophys Res Lett doi:10.1029/2012GL053421.

Peterson, L.C., 2000. Rapid Changes in the Hydrologic Cycle of the Tropical Atlantic During the Last Glacial. Science 290, 1947–1951. doi:10.1126/science.290.5498.1947.

Pope, V.D., Gallani, M.L., Rowntree, P.R., Stratton, R.a., 2000. The impact of new physical parametrizations in the Hadley Centre climate model: HadAM3. Climate Dynamics 16, 123–146. doi:10.1007/s003820050009.

Rhodes, R.H., Brook, E.J., Chiang, J.C., Blunier, T., Maselli, O.J., McConnell, J.R., Romanini, D., Severinghaus, J.P., 2015. Enhanced tropical methane production in response to iceberg discharge in the North Atlantic. Science 348, 1016–1019. doi:10.1126/science.1262005.

Roberts, W.H.G., Valdes, P.J., Payne, A.J., 2014. A new constraint on the size of Heinrich Events from an iceberg/sediment model. Earth and Planetary Science Letters 386, 1–9. doi:10.1016/j.epsl.2013.10.020.

Roberts, W.H.G., Valdes, P.J., Singarayer, J.S., 2017. Can energy fluxes be used to interpret glacial/interglacial precipitation changes in the tropics? Geophysical Research Letters doi:10.1002/2017GL073103.

Schneider, T., 2017. Feedback of Atmosphere-Ocean Coupling on Shifts of the Intertropical Convergence Zone. Geophysical Research Letters 44, 11,644–11,653. doi:10.1002/2017GL075817.

Schneider, T., Bischoff, T., Haug, G.H., 2014. Migrations and dynamics of the intertropical convergence zone. Nature 513, 45–53. doi:10.1038/nature13636.

Sidorenko, D., Rackow, T., Jung, T., Semmler, T., Barbi, D., Danilov, S., Dethloff, K., Dorn, W., Fieg, K., Goessling, H.F., Handorf, D., Harig, S., Hiller, W., Juricke, S., Losch, M., Schröter, J., Sein, D.V., Wang, Q., 2014. Towards multi-resolution global climate modeling with ECHAM6–FESOM. Part I: model formulation and mean climate. Climate Dynamics 44, 757–780. doi:10.1007/s00382-014-2290-6.

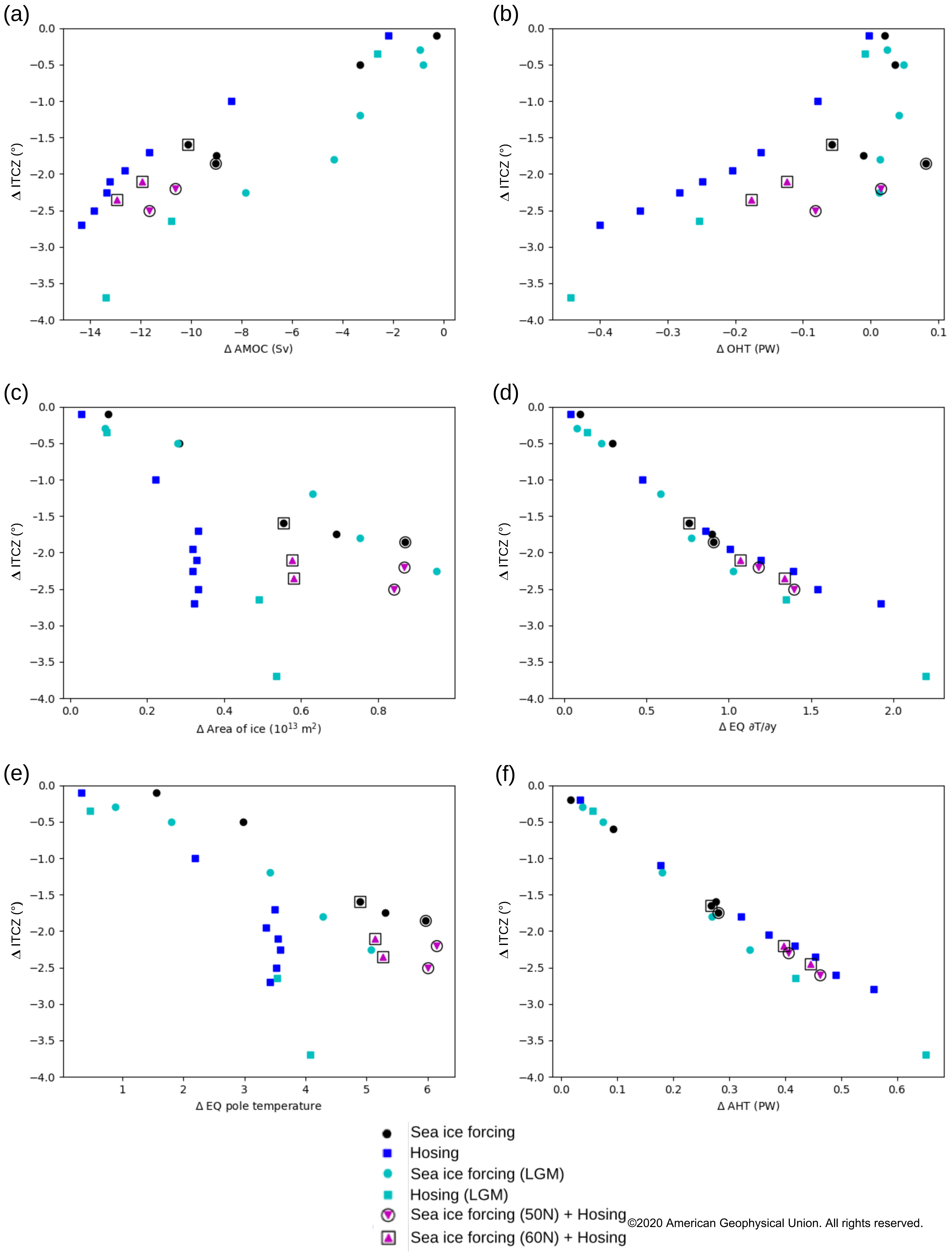
Singarayer, J.S., Valdes, P.J., Roberts, W.H.G., 2017. Ocean dominated expansion and contraction of the late Quaternary tropical rainbelt. Scientific Reports , 1–9doi:10.1038/s41598-017-09816-8.

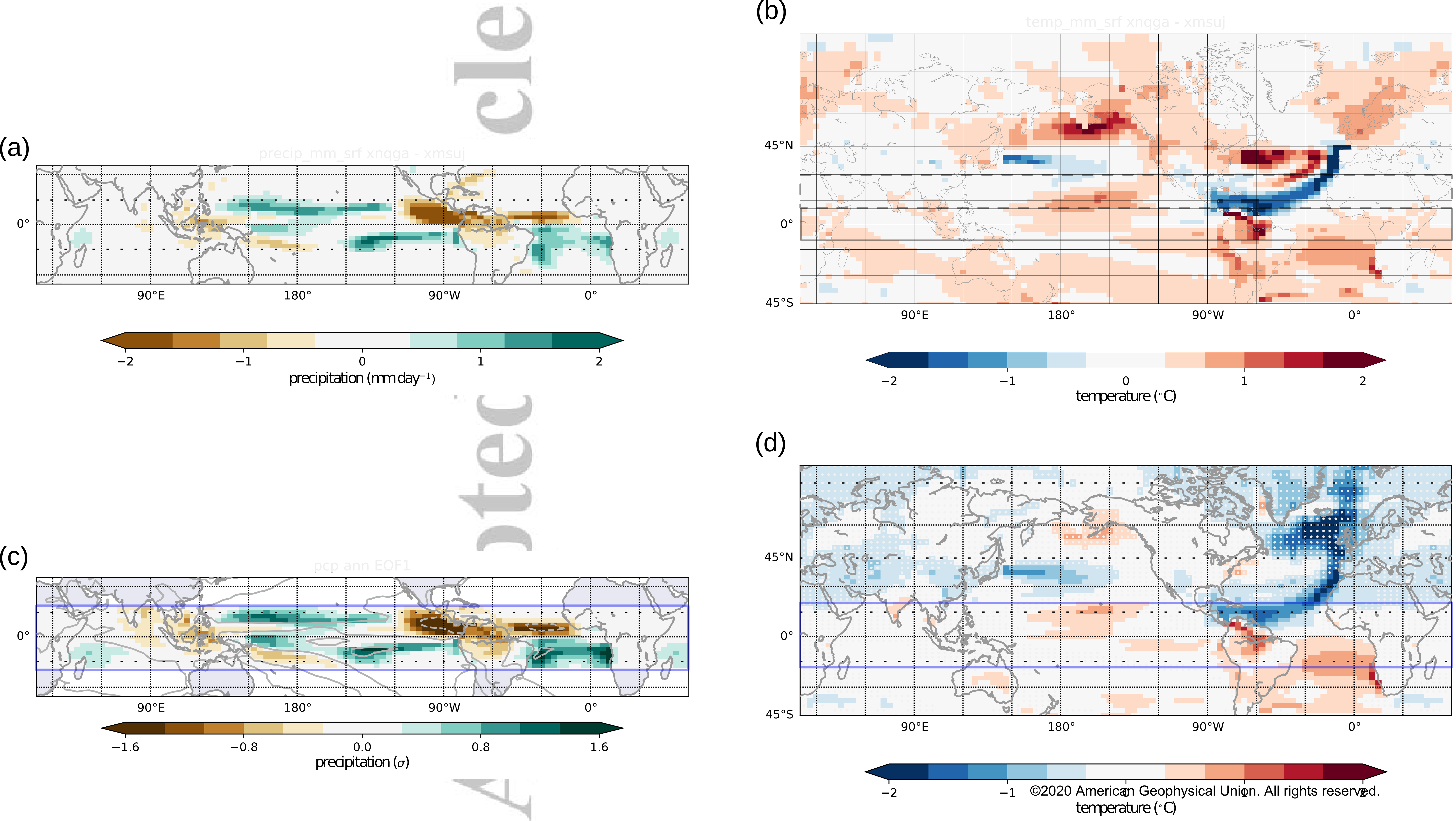
Them, T.R., Schmidt, M.W., Lynch-stieglitz, J., 2015. Millennial-scale tropical atmospheric and Atlantic Ocean circulation change from the Last Glacial Maximum and Marine Isotope Stage 3. Earth and Planetary Science Letters 427, 47–56. doi:10.1016/j.epsl.2015.06.062.

Thornalley, D.J.R., Barker, S., Becker, J., Hall, I.R., Knorr, G., 2013. Abrupt changes in deep Atlantic circulation during the transition to full glacial conditions. Paleoceanography 28, 253–262. doi:10.1002/palo.20025.

Valdes, P.J., Armstrong, E., Badger, M.P., Bradshaw, C.D., Bragg, F., Crucifix, M., Davies-Barnard, T., Day, J., Farnsworth, A., Gordon, C., Hopcroft, P.O., Kennedy, A.T., Lord, N.S., Lunt, D.J., Marzocchi, A., Parry, L.M., Pope, V., Roberts, W.H., Stone, E.J., Tourte, G.J., Williams, J.H., 2017. The BRIDGE HadCM3 family of climate models: HadCM3@Bristol v1.0. Geoscientific Model Development 10, 3715–3743. doi:10.5194/gmd-10-3715-2017.

Vettoretti, G., Peltier, W.R., 2018. Fast physics and slow physics in the nonlinear Dansgaard-Oeschger relaxation oscillation. Journal of Climate 31, 3423–3449. doi:10.1175/JCLI-D-17-0559.1.


(a)
(b)
(c)
(d)
(e)
(f)
Δ ITCZ (°)
Δ AMOC (Sv)
Δ OHT (PW)
Δ Area of ice (10^{13} m^2)
Δ EQ ∂T/∂y
Δ EQ pole temperature
Δ AHT (PW)
Sea ice forcing
Hosing
Sea ice forcing (LGM)
Hosing (LGM)
Sea ice forcing (50N) + Hosing
Sea ice forcing (60N) + Hosing


(a)
precip_mm_srf xnqga - xmsuj
0°
90°E
180°
90°W
0°
−2
−1
0
1
2
precipitation (mm day⁻¹)
(b)
temp_mm_srf xnqga - xmsuj
45°N
0°
45°S
90°E
180°
90°W
0°
−2
−1
0
1
2
temperature (°C)
(c)
pcp ann EOF1
0°
90°E
180°
90°W
0°
−1.6
−0.8
0.0
0.8
1.6
precipitation (σ)
(d)
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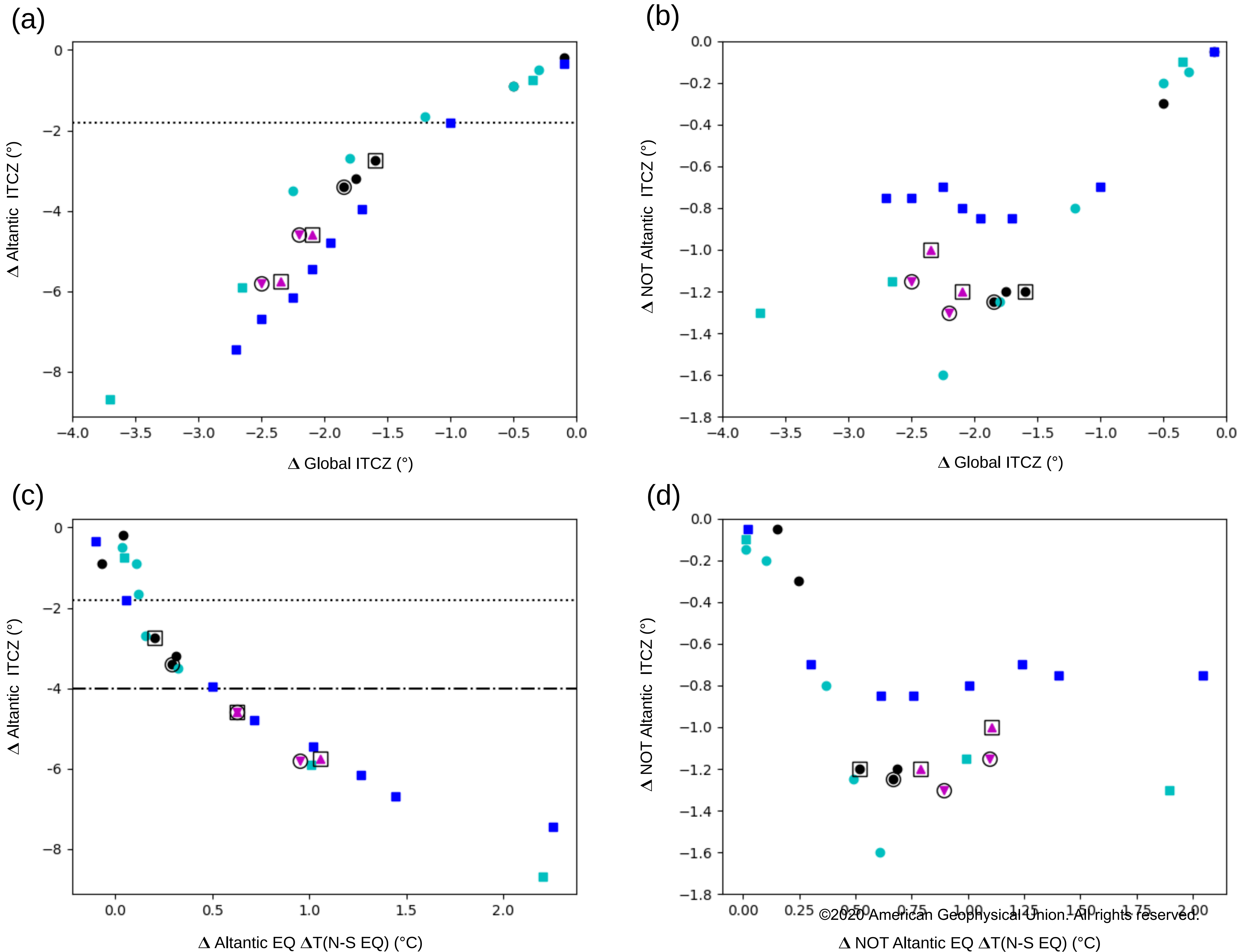
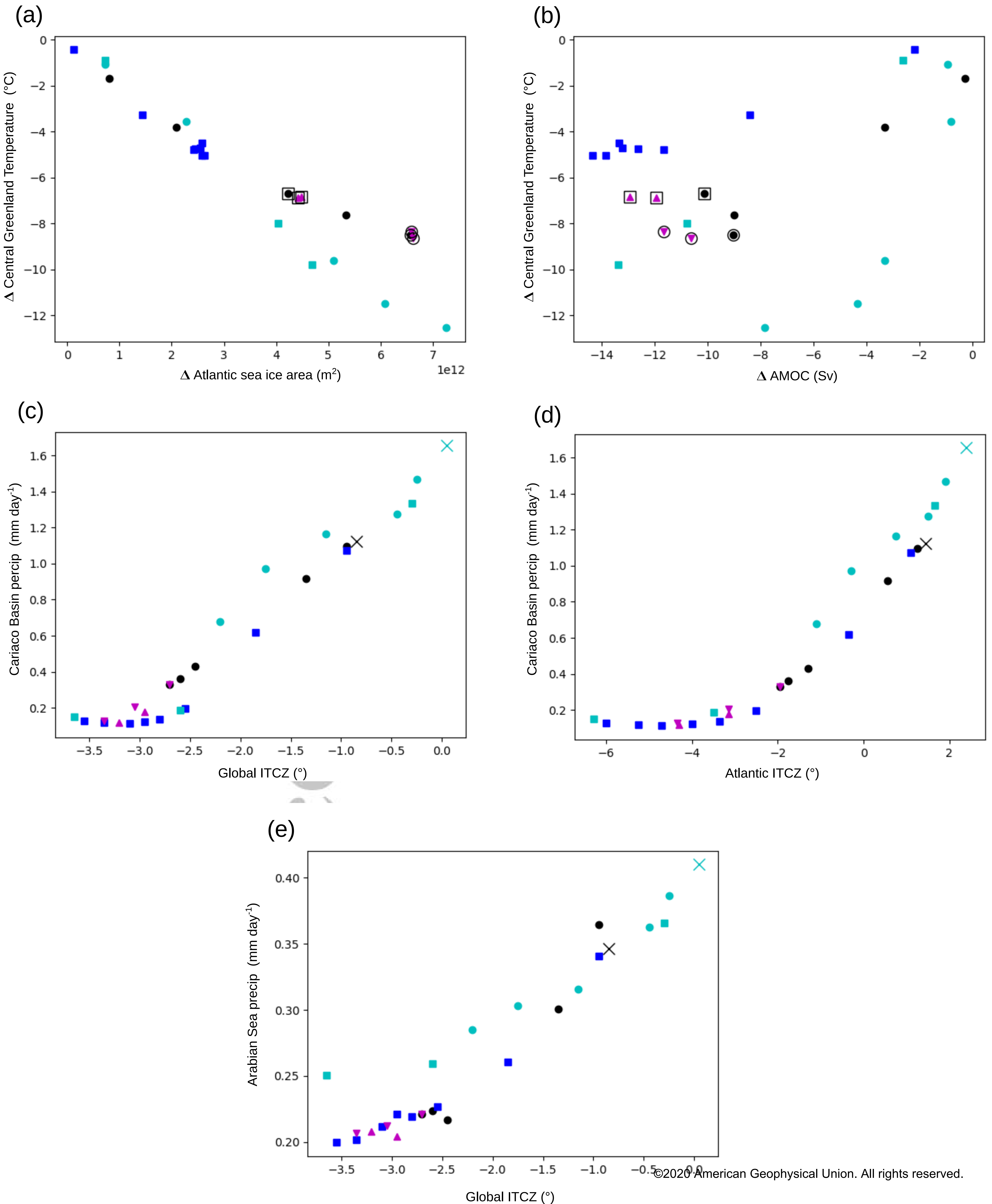
temperature (°C)


(a)
Δ Altantic ITCZ (°)
0
−2
−4
−6
−8
−4.0 −3.5 −3.0 −2.5 −2.0 −1.5 −1.0 −0.5 0.0
Δ Global ITCZ (°)
(b)
Δ NOT Altantic ITCZ (°)
0.0 −0.2 −0.4 −0.6 −0.8 −1.0 −1.2 −1.4 −1.6 −1.8
−4.0 −3.5 −3.0 −2.5 −2.0 −1.5 −1.0 −0.5 0.0
Δ Global ITCZ (°)
(c)
Δ Altantic ITCZ (°)
0
−2
-4
−6
−8
0.0 0.5 1.0 1.5 2.0
Δ Altantic EQ ΔT(N-S EQ) (°C)
(d)
Δ NOT Altantic ITCZ (°)
0.0 −0.2 −0.4 −0.6 −0.8 −1.0 −1.2 −1.4 −1.6 −1.8
0.00 0.25 0.50 0.75 1.00 1.25 1.50 1.75 2.00
Δ NOT Altantic EQ ΔT(N-S EQ) (°C)


(a)
(b)
(c)
(d)
(e)
Δ Central Greenland Temperature (°C)
Δ Atlantic sea ice area (m^2)
1e12
Δ AMOC (Sv)
Cariaco Basin percip (mm day^{-1})
Global ITCZ (°)
Atlantic ITCZ (°)
Arabian Sea precip (mm day^{-1})